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MORPHOLOGICAL CONTROLS ON THE DOWNSTREAM PASSAGE OF A SEDIMENT WAVE IN A GRAVEL-BED STREAM

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ABSTRACT

Sediment waves in river systems have been widely reported, although few studies have examined the interaction between these waves and the morphology of the reaches through which they pass. This interaction determines how waves are modified as they propagate downstream. This study documents the origin and downstream passage of an avulsion-generated sediment wave through a 374 m study reach of the Allt Dubhaig, Scotland. A nested survey framework was adopted, with volumes calculated from cross-sections spaced between 10 and 40 m apart documenting the origin and downstream passage of the wave. The wave moved through an intensively (*c*. 1 m cross-section spacing) monitored 120 m stretch (Reach A) within the study reach, allowing assessment of sediment exchanges between the incoming wave and the local morphology. Successive surveys show the movement of the wave through and out of the reach, and also that areas where wave sediment was deposited did not always correspond with areas of subsequent erosion. Reach A was divided into three morphologically distinct sub-reaches (1A, 2A and 3A) within which sediment fluxes and the three-dimensional distribution of erosion/deposition were estimated. Sediment wave input into 1A and 2A (relatively stable sub-reaches) caused forced bar aggradation and erosion of sediment from elsewhere within the reach, which then became part of the wave. The downstream transfer of this sediment into unstable 3A caused aggradation and, in response, widespread erosion which increased the magnitude of the sediment wave as it exited reach A. Sediment exchange between the recipient reach and the wave depends upon local morphological stability and is a crucial process affecting wave magnitude and attenuation. The macroscale sediment wave interacted with, rather than overwhelmed, the recipient morphology. © 1998 John Wiley & Sons, Ltd.

KEY WORDS: sediment wave; gravel-bed river; sediment storage; channel morphology

INTRODUCTION

Sediment transfer through river systems commonly occurs in the form of bed waves (or slugs) which migrate downstream (Church and Jones, 1982; Hoey, 1992; Nicholas *et al*., 1995). These features have been identified at a range of prototype scales, from individual bedforms (mesoforms, wavelength 10^{-1}–10^{2} m, scale with flow depth) to unit or complex bars (macroforms, wavelength 10^{1}–10^{3} m, scale with channel width), and finally to bar assemblages (megaforms, wavelength >10^{3} m, scale with channel width or drainage basin morphology). The morphological expression of bed waves varies with their scale as do the nature and magnitude of their impact on downstream reaches. Bed waves can be generated within river channels in the absence of changes in the rate of external sediment supply from hillslopes or tributaries (e.g. Griffiths, 1979; Meade, 1985; Kuhnle and Southard, 1988; Ashmore, 1991). Such waves (‘endogenous’ waves or ‘autopulses’) have usually been reported for the smaller scale features, although they have been produced at relatively larger scales in flume experiments (e.g. Ashmore, 1991; Hoey and Sutherland, 1991; Warburton and Davies, 1994). Wave generation where external sediment supply is accelerated (‘exogenous’ waves or ‘allopulses’) has been noted in cases of natural sediment

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supply events (e.g. Beschta, 1983) and where human activities accelerate sediment input (e.g. Gilbert, 1917; Pickup *et al*., 1983; Knighton, 1989; Madej and Ozaki, 1996).

The most detailed studies of sediment wave movement have been conducted with regard to the transfer of mining waste through river systems as megaforms (e.g. Gilbert, 1917; Pickup *et al*., 1983; James, 1989; Knighton, 1989; Macklin and Lewin, 1989). In these cases the sediment delivery systems are severely disturbed, often by sediment of different calibre from the 'natural' sediment in the system at that location. The models of disturbance and recovery of these systems (e.g. James, 1989; Knighton, 1989) suggest limited interaction between the wave sediment and the existing bed material. However, where the waves involve relatively small amounts of sediment, little is known about how the wave sediment interacts with sediment stored within the receiving river reach(es).

The dispersion of tracer gravels has been related to reach morphology (e.g. Hassan *et al*., 1991) and it is suggested that reach response to an incoming bed wave will depend on reach morphology (Macklin and Lewin, 1989) and on the amount and distribution of stored sediment within the reach. This implies that recent hydrological and sediment supply histories will effect reach response (Ferguson and Werritty, 1983; Hoey, 1994). Limited evidence from tracers suggests that dispersion is rapid and that incoming sediment quickly mixes with that in the receiving reach (Mosely, 1978; Kondolf and Matthews, 1986). However, these studies do not assess the relationship between disperson of wave sediment and the attenuation of sediment wave form which is usually assumed to occur during downstream transfer (Nicholas *et al*., 1995). This is a key issue for the understanding of wave dynamics and needs to be addressed before progress can be made in the modelling of wave movement.

This paper describes the downstream passage of a macroscale endogenous sediment wave and monitors its interaction with local morphology as it passes through a 120 m reach. Wave sediment is of comparable calibre to reach bed material. The specific aims of this study are to: (1) assess the significance of local morphology upon the passage of the wave; (2) determine the extent of sediment exchange between the recipient reach and the wave, and the effect of this upon wave magnitude; and (3) describe channel response to wave input and assess the transience of these changes.

FIELDSITE

The Allt Dubhaig is a headwater tributary of the River Tay in the Scottish Highlands described by Ferguson and Ashworth (1991) and Ferguson *et al*. (1996). The 3·5 km long alluvial channel, which begins where the river emerges from an area of moraine and crosses a bedrock bar, has a strongly concave long profile. Cross-sections used in this study are numbered according to their distance (in metres) downstream of this point.

The study reach (Figure 1) is between surveyed sections at 845 m and 1219 m. Reach average channel slope is 0·011 m/m, channel pattern is transitional wandering/meandering, and average surface D_{50} (based upon bulk samples of 941 and 420 kg respectively, truncated at 2 mm from sections 845 m and 1062 m) is 54 mm. This reach has been monitored since November 1990 and has exhibited frequent bar and channel migration but minimal bank erosion. There are several abandoned channels in the reach, all of which receive a small amount of flow at bankfull conditions. Stage was recorded every 15 min at gauging station Q3, located between sections 1042 and 1062 m.

Part of this reach (Reach A), located between 1000 m and 1131 m, forms the basis for detailed data collection (Figure 2). The upstream two-thirds of the reach are dominated by two 'forced' bars (in the terminology of Seminara and Tubino, 1989) which remained stable during the first half of the study (reach A has been monitored since September 1991) with adjacent pool/riffle units being subject to minor alteration. The channel has average bankfull width:depth ratio of approximately 10, and the lack of appreciable morphological change suggests that this part of Reach A behaved as a 'transfer' reach (e.g. Church and Jones, 1982) even though the bars represent considerable sediment stores. These constitute an inactive sediment store in the terminology of Hoey and Sutherland (1991) and Hoey (1996), which have the potential to be re-activated by either active erosion (dynamic sediment transfer) or as a result of reach aggradation (static sediment transfer). The downstream one-third of the reach is less stable and is dominated by more transient 'free' bar storage which was continually modified during the first half of the study (1990–93). High width:depth ratios (average

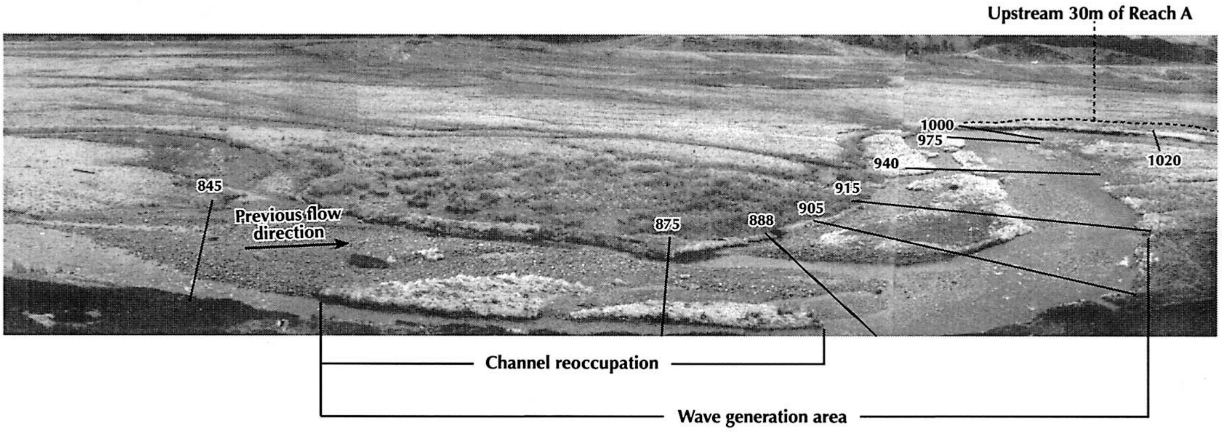

Figure 1. Upstream half of the study reach (830–1030 m) including the wave generation area, February 1993, discharge = 1 m^3 s^{-1}. The sediment wave was generated and transported into Reach A during the January 1993 flood. Previously, flow was confined to the left and dissected the head of the medial bar between sections 845 and 888. Cross-section numbers refer to distance from the start of the alluvial channel and may be used as an approximate scale

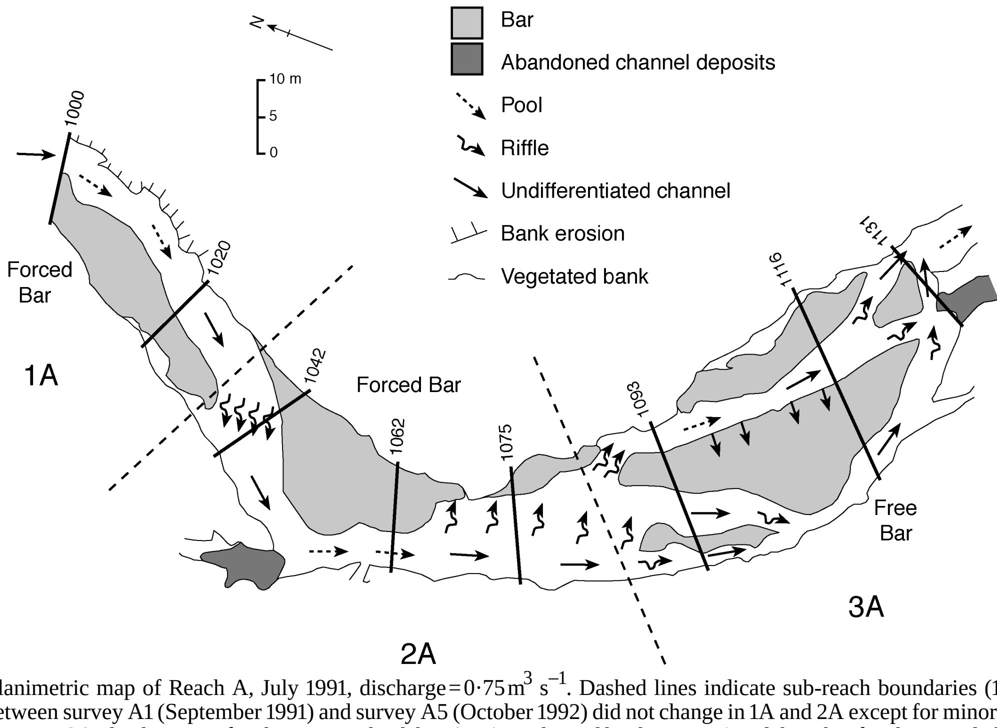

Figure 2. Planimetric map of Reach A, July 1991, discharge = 0·75 m^3 s^{-1}. Dashed lines indicate sub-reach boundaries (1A, 2A, 3A). Planform between survey A1 (September 1991) and survey A5 (October 1992) did not change in 1A and 2A except for minor migration of pool/riffle units. In 3A, the dominant free bar was cut back by 10 m (as indicated by the arrows) and the other free bars eroded or migrated. The inactive channel next to the right bank in 3A (approximately 0·05 per cent of total low flow) was blocked by deposition at the tail of the riffle on the 2A/3A boundary in January 1992 and did not reactivate. Eight of the 18 cross-sections covering the study reach are located in Reach A (sections 1000–1131)

approximately 47) and considerable sediment storage indicate that this part of Reach A is a sedimentation zone or storage reach (e.g. Church and Jones, 1982; Church, 1983) dominated by 'semi-active' sediment storage (Hoey and Sutherland, 1991). Ongoing bank erosion in Reach A only affects the >1·5 m left bank from 1000 m to 1030 m, with the lower banks elsewhere being stable and vegetated.

Table I. Cross-section survey completion dates within the study reach and Reach A

Study reach (845–1219 m)		Reach A (1000–1131 m)		Number of flow events prior to survey
Survey no.	Date	Survey no.	Date	
1	21 November 1990			–
2	31 May 1991			6†‡
		A1	21 September 1991	1
		A2	17 January 1992	20
		A3	20 March 1992	7
3	25 May 1992			4
		A4	3 September 1992	6
		A5	9 October 1992	7
		A6	15 February 1993*	13
4	1 March 1993*			2
5	18 May 1993			2
		A7	18 July 1993	2
6	30 November 1994			32†
7	22 June 1995			17†
		A8	18 September 1995	0†

* The first survey after the sediment wave was generated
Number of flow events is the number of events in which the threshold stage for bedload transport was exceeded
† Estimated from gauging station 1 km upstream
‡ One month of record missing

DATA COLLECTION AND ANALYSIS

A total of 18 cross-sections located at pool and riffle sites within the study reach were surveyed once or twice per year (Table I; surveys 1–7). These were used to identify the zone where the sediment wave was generated and to monitor its downstream passage. Volumetric change between surveys for successive cross-sections was determined using the prism formula as described by Ferguson and Ashworth (1992). This formula is more accurate where the section spacing is small so that morphological changes between sections are quantitatively represented by changes at the sections. Section spacing (mean *c*. 17 m) is too great to obtain reliable topographic data (Lane *et al*., 1994), but these data may be used to identify general changes and relative downstream flux rates through the study reach.

Sediment storage was monitored in Reach A between September 1991 and September 1995 using a network of 115 cross-sections spaced between 1 and 1·5 m apart (Table I; surveys A1–A8). This dense survey interval (cross-stream points were no more than 1 m apart) ensures accurate quantification of bed topography (Lane *et al*., 1994; Ashmore and Church, in press). Lane *et al*. (1994) suggested that errors increase rapidly if cross-section spacing exceeds 3 m for channels of similar gradient to the Allt Dubhaig between 10 and 20 m wide. The dense spatial scale prevented full surveys being carried out after every bedload transporting event; instead, convenient 3 day periods of low flow were used, these being scheduled after one or more event above the threshold for bedload transport had occurred. There were seven full surveys between September 1991 and May 1993, and a further resurvey in September 1995 (Table I).

Cross-section data from Reach A were used to construct a digital elevation model (DEM) of bed topography. All cross-section points are expressed relative to a sloping datum corresponding with mean reach slope. Mean bed and bankfull elevations decrease approximately linearly through the reach, and use of a sloping datum ensures that a given volume of sediment has the same impact on the sediment storage calculations wherever it is located within the reach. This datum was subtracted from the elevation of each point to transform the data relative to a horizontal plane. The data from each survey were interpolated to a 1 m×1 m regular grid and contoured (using UNIMAP) to define the reach surface topography. The interpolation method was bilinear with quadratic interpolation used to achieve smoothing. The spatial distribution of erosion and deposition was determined by subtraction of z_2 (elevation interpolated from data at $t=2$) from z_1 (interpolated at $t=1$ at exactly the same grid locations as z_2). The resultant net elevation change at each node $Z_2=z_1-z_2$ was contoured to reveal the extent of erosion and deposition between surveys.

Reach A cross-sections were also used to determine the downstream distribution of sediment fluxes and to develop a within-reach sediment budget. The reach was divided into 38 segments with boundaries located after

every third cross-section and marginal boundaries defined by bank foot location. The volume of sediment stored was calculated by summation of the interpolated elevations, z_1, at each node within a segment, which was then divided by the number of nodes per 1 m^2 (usually $10\times10/\mathrm{m}^2$).

Net volumetric change between successive surveys was calculated by subtraction. All segment boundaries were constant, except for segments 1–10 where bank erosion moved the periphery by up to 1·5 m during some floods. For consistency, volumetric changes were calculated relative to the boundary at the earlier survey; for example, net change between surveys A7 and A8 was calculated using the segment boundary from survey A7. This method excludes the bank sediment – nearly all of which is fine grained and does not contribute to the gravel sediment budget – from the volumetric calculation therefore avoiding spuriously high erosion figures. These data allow a simple within-reach sediment budget to be determined assuming sediment continuity where the inputs (I), outputs (O) and changes in sediment storage (ΔS) balance (Griffiths, 1979; Dietrich *et al*., 1982; McLean, 1990; Ferguson and Ashworth, 1992). The sediment input passing through the upstream reach boundary was set to the minimum value which ensured non- negative downstream output between all sections (e.g. Ferguson and Ashworth, 1992; Goff and Ashmore, 1994), and all flux data must therefore be considered to be minima. This approach was also applied to net cross-section change data for all adjacent sections from 845 m and 1219 m to determine downstream fluxes through the study reach.

The use of a gridded DEM offers two main advantages over other methods of volume calculation from cross-sections, such as the prism formula: (1) it is more accurate as interpolation replicates the bed surface on the basis of all points in the vicinity; and (2) it is computationally faster. However, there are problems: (1) interpolation is only as good as the data upon which it is based – microscale topography between sections affects accuracy (this applies to all methods of volume calculation from cross-sections), but these errors are minimized with a small survey spacing; and (2) a square grid was used. For an irregular area containing incomplete cells the volume algorithm uses half or three-quarters of z_1 depending upon the proportion of the cell falling within the area. Sensitivity testing revealed that this error is minimal, although it is a function of the interpolated grid size.

The prism formula-derived data (study reach, sections 845–1219) are less accurate than the DEM data (Reach A) since interpolation is taking place over longer distances, therefore direct comparison between volumetric flux data calculated using these two approaches is not reliable. Errors reflect both the spacing of surveyed sections and the position of the sections relative to reach morphology. However, calculation errors in the data from both methods are small in comparison to the magnitude of the fluxes (Ashmore and Church, in press), so that conclusions derived from these data are valid.

SEDIMENT WAVE GENERATION AND DOWNSTREAM MIGRATION

Snowmelt and rain-on-snow in January 1993 resulted in an unusually large and long duration multiple peak flood. Ice and snow during this period confined overbank flow and prevented application of the rating equation. Therefore the calculated discharge of 20 m^3 s^{-1} must be viewed as an estimate (bankfull discharge is approximately 9 m^3 s^{-1}). The duration of the flood above a bedload threshold (defined as the minimum flow required to transport gravel into a row of bedload traps located 1 km downstream; see Wathen *et al*. (1995) for details) was 35 h. Despite the impact of the snow and ice this is clearly the largest and longest flood event recorded at Q3 (1991–96). Considerable channel change in response to this extreme event occurred along the upstream 80 m of the main study reach releasing a macroscale sediment wave. The first field visit after the flood (20/1/93) indicated an avulsion between sections 845 and 888. Prior to this (surveys 1–3, A1–A5, Table I), the channel flowed towards the left bank dissecting the head of the partly vegetated medial bar (Figure 1). The flood caused aggradation of up to 0·35 m on the bar head resulting in reoccupation of the 40 m long abandoned channel to the right where the bed was lowered by up to 0·5 m (Figure 3). Channel reoccupation also modified downstream flow fields which resulted in partial erosion of a point bar between sections 905 and 915.

A sediment wave is defined as ‘an increase in the amount of sediment stored in a particular channel reach relative either to that reach at different times, or to the adjacent upstream and downstream reaches at a given time’ (Hoey and Sutherland, 1991; Hoey, 1992). As such, the form of the waves may be stepped rather than smooth, and they may not be as immediately obvious in three dimensions in the field as they appear in collapsed

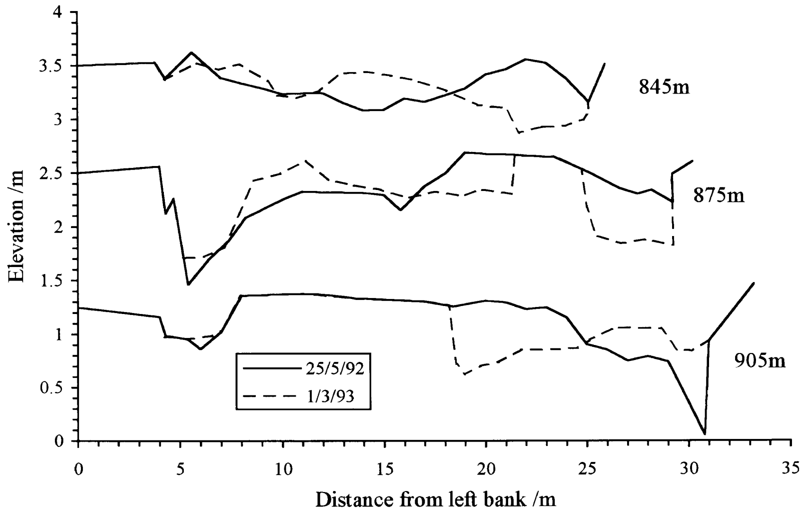

Figure 3. Pre-wave (survey 3, May 1992) and post-wave (survey 4, March 1993) surveys of sections 845, 875 and 905 with flow into the page. Section elevations are for plotting purposes and are not tied into a datum. Distances are expressed relative to the left bank. Note erosion associated with channel reoccupation adjacent to the right bank of sections 845 and 875 and erosion of a point bar between 18 and 25 m along section 905

one-dimensional graphical representations. Bed elevation changes at sections from 845 to 1219, surveys 3 and 4, indicates substantial upstream erosion (845 to 940) and downstream deposition (940 to 1219) consistent with the preceding definition. When volumes between sections are calculated from which volumetric fluxes are derived, these show a clear waveform with higher fluxes in the areas of upstream erosion, which decline rapidly as the sediment is deposited downstream (e.g. Hoey and Sutherland, 1991). The minimum peak wave output from section 940 was 256 m^3 (this figure must be viewed as an estimate given the section spacing), 220 m^3 of which was deposited before section 1062 (Figure 4a). Subsequent surveys (4 and 5) reveal no further wave-like generation of sediment from the source, but downstream migration is apparent with erosion between sections 1042 and 1062 (Figure 4b). Later surveys of the study reach are incomplete and prevent tracking of the wave downstream of Reach A. The spatial pattern of sediment flux reflects that of aggradation/degradation (Figure 4). However, the absolute flux values can be misleading: between surveys 3 and 4 (Figure 4a) there were 28 flow events exceeding the threshold stage for bedload transport, lasting for *c.* 20 days in total, whereas between surveys 4 and 5 there were two events lasting *c.* 1·5 days. The total bedload transport in the former interval would have far exceeded that in the latter, but both are assumed to have a minimum flux between sections of O (Figure 4). The fluxes in the earlier period (Figure 4a) are thus underestimated more severely than those in the latter (Figure 4b) and should be increased by an unknown constant amount.

The data for the entire reach indicate the presence of a sediment wave but the data resolution is too broad to allow examination of its behaviour. However, the detailed morphological data from reach A allow description of the passage of this wave through this reach and, in particular, morphological explanations for wave form and magnitude. The wave was generated by an extreme event which was not repeated during the study. Subsequent surveys monitor the passage of the wave in response to a succession of smaller (up to and including bankful, 9 $m^3\ s^{-1}$) events. Net change for each segment was determined for successive time periods (surveys A5–A6, A6–A7 and A7–A8). Note that the first time period includes three months prior to wave generation (and two bankfull events). Net change data indicate upstream aggradation (segments 1–23) and some downstream erosion as the wave enters the reach (Figure 5a). Subsequently, sediment is eroded from upstream locations and there is deposition towards the end of the reach between surveys A6 and A7. Deposition in these downstream locations is followed by significant erosion (in excess of previous deposition) between surveys A7 and A8 (Figure 5b). Close inspection of the one-dimensional data presented in Figures 5a and 5b reveals that areas of deposition do not perfectly match areas of subsequent erosion and therefore decreasing sediment fluxes (erosion) from the same location (Figure 5c). In addition, between surveys A6 and A8 more sediment exits the

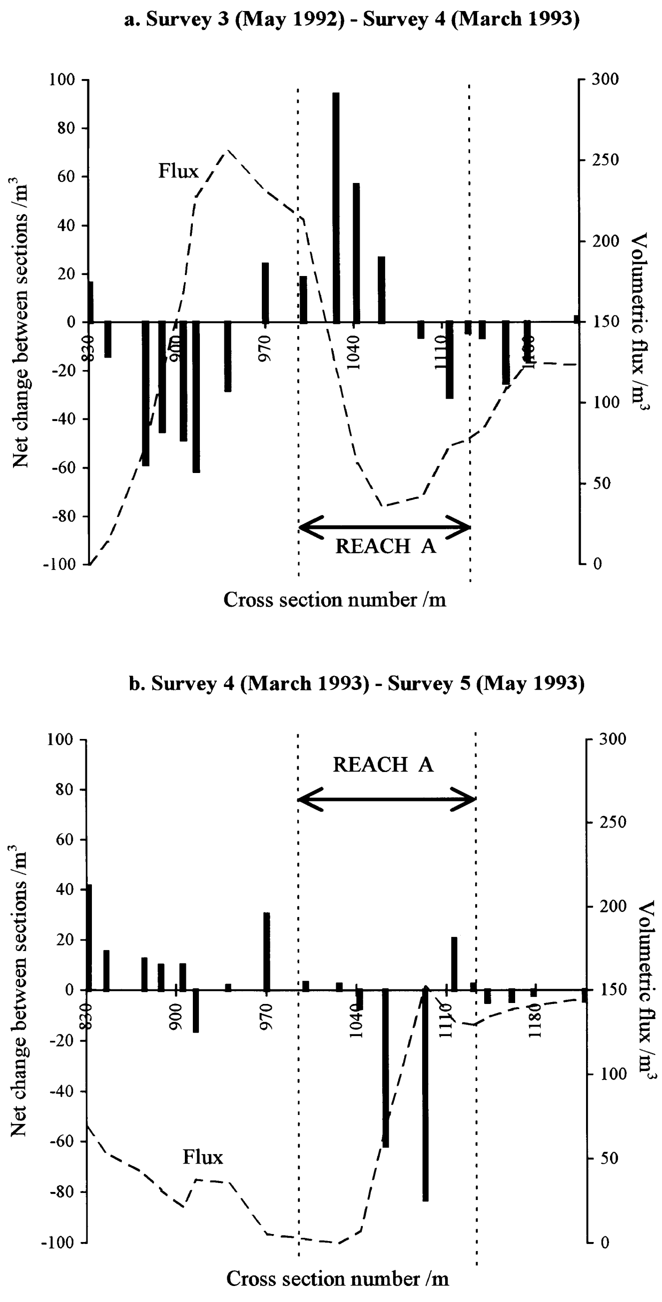

Figure 4. Net volumetric change and downstream sediment flux between adjacent cross-sections (m^3), survey 3 (May 1992) to survey 5 (May 1993). The study reach boundary was extended to include sections 804 and 830 which aggraded during each survey. This allowed more accurate estimation of the sediment flux into the reach which was set to the minimum value to ensure non-negative transfer between all sections. Dotted lines indicate the location of Reach A. Note the downstream passage of a waveform (a) and subsequent erosion within Reach A (b)

reach than enters it. These data indicate that as the wave passes through the reach, some sediment is eroded from areas which were previously not subject to wave deposition or that erosion exceeds previous wave deposition. The wave form is therefore preserved as a result of sediment exchange, and not all of the original wave sediment

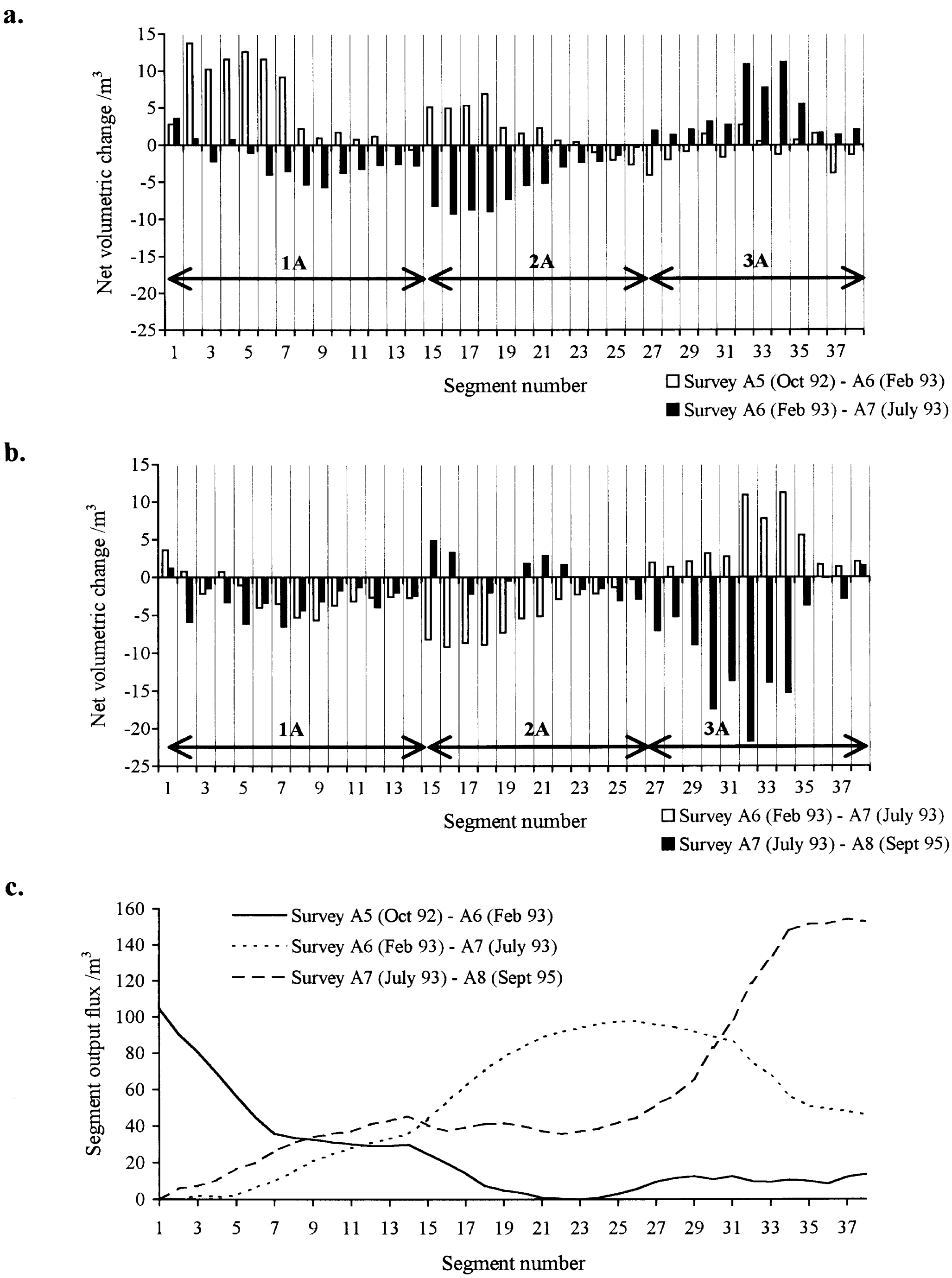

Figure 5. Net volumetric change within Reach A segments and reach sediment fluxes, survey A5 (October 1992) to survey A8 (September 1995). Note the downstream shift in the cycle of deposition and erosion as the wave passes through the reach and the imperfect correspondence between areas of deposition and subsequent erosion. Sediment fluxes indicate areas of deposition (decreasing flux) and erosion (increasing flux) consistent with the passage of a waveform

remains within it. Explanations for this must be sought with reference to the impact of the wave upon recipient morphology and vice versa.

Table II. Minimum Reach A input and output during the study period

Reach A survey date	Survey numbers	Volumetric flux (m^3)		
		Min. input	Min. output	Net change
October 1992–February 1993	A5–A6	108	14	+94
February 1993–July 1993	A6–A7	4	46	–42
July 1993–September 1995	A7–A8	1	153	–152
Wave study period	A5–A8	113	213	–100

Input is the value required to ensure non-negative transfer throughout the reach. Note that the wave increased sediment output as it migrated downstream

IMPACT OF THE WAVE ON THE REACH

Reach input and output fluxes demonstrate the magnitude and timing of the sediment wave (Table II). The minimum volume of the wave as it entered the reach (derived from segment data) was 108 m^3, with <6 m^3 (minimum) entering between survey A6 and the end of the study. Minimum output from the reach over the subsequent 31 months (surveys A6 to A8) totalled 199 m^3, and the net change in Reach A storage during this period was a deficit of 100 m^3. These results suggest that disturbances induced by the wave resulted in a release of sediment which increased the total volume of sediment in the wave as it moved through the reach.

Breaking the reach down into sub-reaches allows closer examination of the effect of reach morphology on wave transfer. Reach A is divided into three sub-reaches, 1A (segments 1–10), 2A (segments 11–22) and 3A (segments 23–38), based upon dominant riffle locations (Figure 2). It is instructive to examine net changes in each sub-reach since the start of the study, as the recent sediment supply history affects reach morphology and stability (e.g. Church and Jones, 1982; Ferguson and Werritty, 1983; Hoey, 1994). Sediment fluxes into and out from these sub-reaches were determined and the cumulative flux (net cumulative volumetric change in each sub-reach since the start of the study, survey 1) was used to identify storage behaviour. Prior to the wave entering the reach, 1A remained stable with only slight aggradation after survey A3 (Figure 6). Storage in 2A declined overall due to periodic riffle and pool migration, and erosion of a low bar-attached platform between surveys A3 and A4, although quantities are small and there was minor aggradation between surveys A4 and A5. In 3A, considerable erosion occurred as a medial bar was eroded, initially rapidly (surveys A1 and A2), then at a lower rate (see Figure 2). These initial conditions are significant and may be summarized as relative stability in 1A and 2A, and degradation and frequent morphological change in 3A; this is consistent with classification of the former as transfer reaches and the latter as a storage reach (Church and Jones, 1982). The three-dimensional impact of the wave upon these sub-reaches after survey A5 is described below.

Sub-reach 1A. Entry of the wave between surveys A5 and A6 was responsible for considerable aggradation on the forced bar (in excess of 0·45 m in places) and progradation towards the left bank together with aggradation in adjacent pools (Figure 7a). Deposition of wave sediment was concentrated in the first 25 m of this sub-reach causing a reduction in upstream channel capacity and bank erosion of up to 1·5 m. Further downstream, the channel incised moderately. Upstream aggradation was followed by minimal changes to bed elevation with more significant erosion confined to the channel in the downstream 10 m of the sub-reach (Figure 7b) where the quantity of previously deposited wave sediment was minimal. There was net aggradation in 1A between the wave entering and the end of the study, as much as the sediment deposited on the forced bars remained in storage (Figure 6).

Sub-reach 2A. Wave entry resulted in up to 0·44 m of bar aggradation (Figure 7c) and, in general, the adjacent channel eroded during the same time period. Flow confinement due to increased bar height was responsible for this erosion which continued during the next series of floods (Figure 7d). The bar margins were also trimmed. In net terms, erosion exceeded previous aggradation (Figure 6) indicating that the wave sediment which was not mobile (no significant changes to bar elevation after aggradation, Figure 7d) triggered the release of new sediment from the adjacent channel which may now be considered to be part of the wave as it passes through the reach. A similar process was responsible for generating the wave at the start of the study reach. Aggradation on

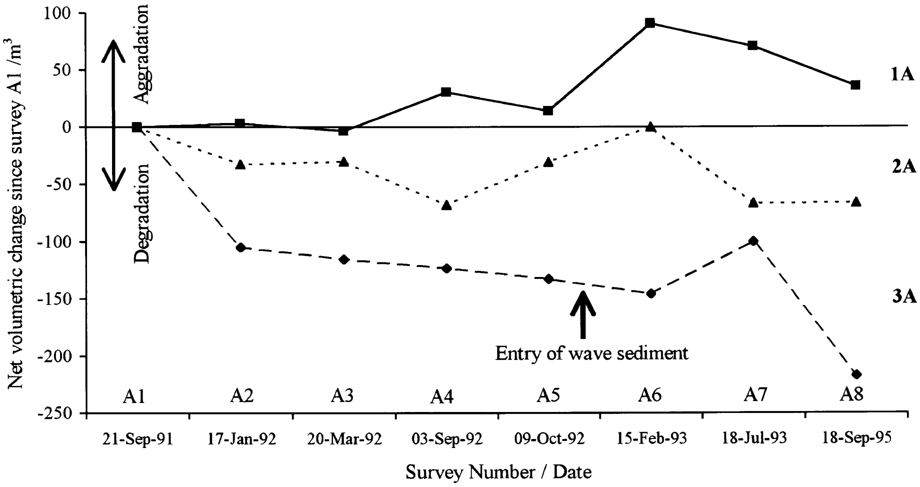

Figure 6. Net volumetric change since survey A1 in sub-reaches 1A, 2A and 3A. Data indicate pre-wave sub-reach characteristics and post-wave response

the bar head at section 845 (Figure 1) created local instability which caused a within-channel avulsion and wave generation.

Sub-reach 3A. Whilst 1A and 2A were responding to the influx of wave sediment, 3A eroded (Figure 8a) as the medial bar (see Figure 2) was cut back. The adjacent channel aggraded by up to 0·29 m during this period (it is not clear whether this was part of the initial wave or sediment released from 1A and 2A due to the entry of the wave). Slight net aggradation (Figure 8b) of 45 m^3 (Figure 6) occurred in 3A between surveys A6 and A7 in response to the wave-induced disturbances in 1A and 2A. It is likely that this large increase in sediment storage caused instability (the reach had previously eroded throughout the study, Figure 6) resulting in net erosion between surveys A7 and A8 (a loss of 117 m^3, Figure 6) when the channel incised towards the left bank and aggraded towards the right (Figure 8c). Initial instability in response to upstream input increased sediment exchange between the sub-reach and the wave and increased the magnitude of the wave as it moved downstream.

The effect of the wave varies between the sub-reaches. Any disturbances in 1A and 2A are transient as the morphology is stable enough to absorb excess sediment with only minor alteration to channel planform. The effect of the wave is more permanent in 3A as this previously unstable reach has responded to increased upstream supply through morphological change as free bars have migrated and/or eroded. Sub-reaches 1A and 2A, although storing large quantities of sediment, are transfer reaches (Church and Jones, 1982) which exchange sediment and modify the wave rather than permitting uninterrupted transfer. Most storage reaches such as 3A absorb incoming waves and may eventually allow the wave sediment to leave, often as a wave of different size (e.g. Church and Jones, 1982; Church, 1983; Macklin and Lewin, 1989; Hoey, 1994). However, if persistent erosion (Figure 6) is assumed to be a response to ‘over-aggradation’ (a hypothezised state where sediment storage exceeds local storage capacity resulting in instability and erosional adjustment (Lane *et al.*, 1996)) then the wave sediment input exaggerated this instability resulting in channel switching and continued passage of the wave. In this case the wave is self-preserving, with its continued existence and downstream passage being responsible for, and derived from, channel disturbance.

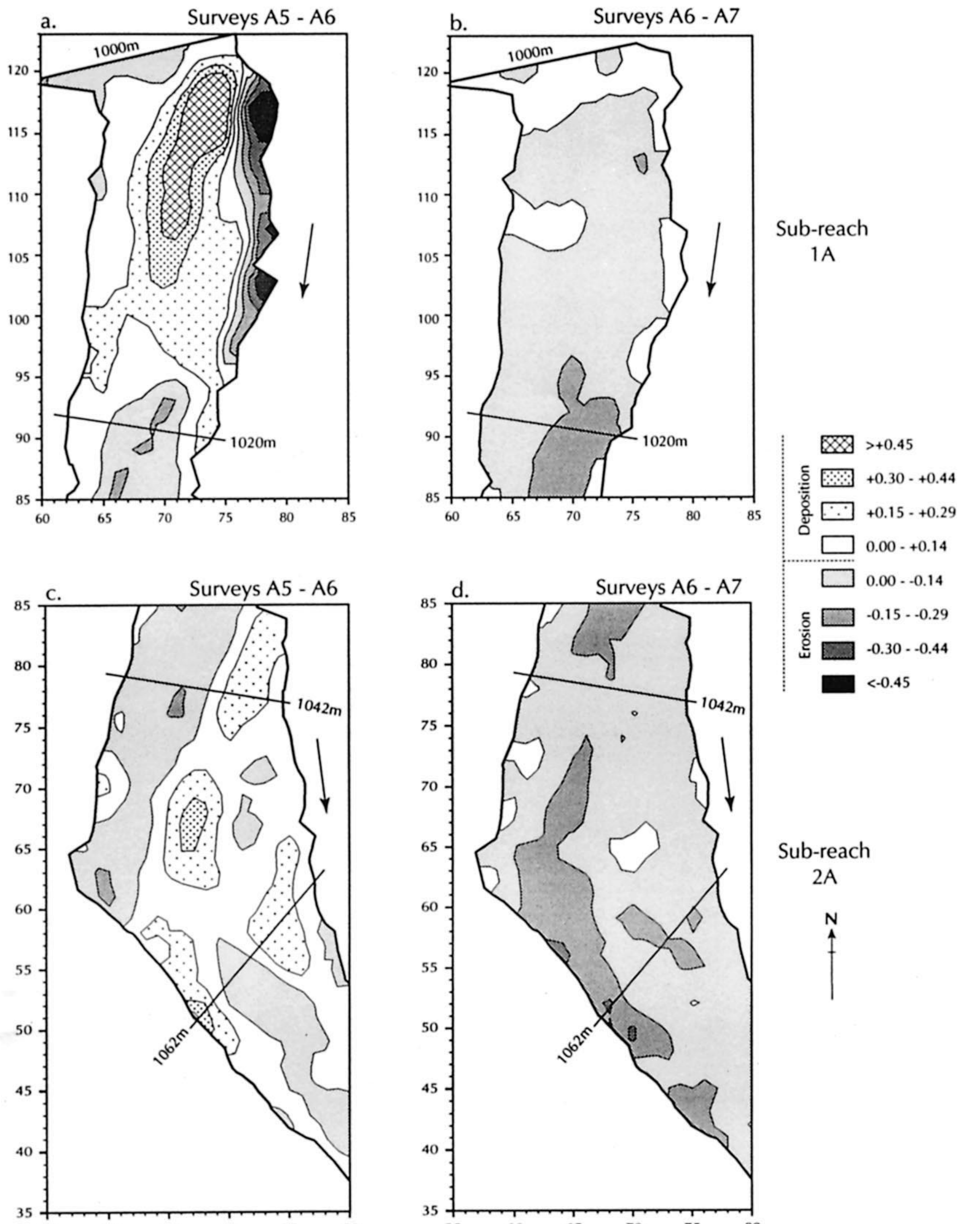

Figure 7. Spatial distribution of deposition (>0 m) and erosion (<0 m) in sub-reaches 1A and 2A between surveys A5 and A6 (a,c) and A6 and A7 (b,d). Note forced bar aggradation due to wave input and subsequent erosion from the adjacent/downstream channel. Sub-reach boundaries are derived from survey A6 (a,c) and A7 (b,d). Units are metres. Labelled cross-sections enable positions to be located with reference to Figure 1 and 2. Arrows show flow direction

DISCUSSION

The above results indicate that macroscale sediment wave form depends upon sediment exchange between the wave and the reach through which it is passing. The magnitude and distribution of this exchange is tempered by reach morphology and morphological stability and therefore the recent history of sediment supply (e.g. Ferguson and Werritty, 1983; Hoey, 1994; Lane *et al*., 1994), including the impact of previous waves (Church and Jones, 1982). The wave in this case is small compared with other reported sediment waves, most of which result from large-scale disturbances such as mining (e.g. Gilbert, 1917; Pickup *et al*., 1983) or channel diversion

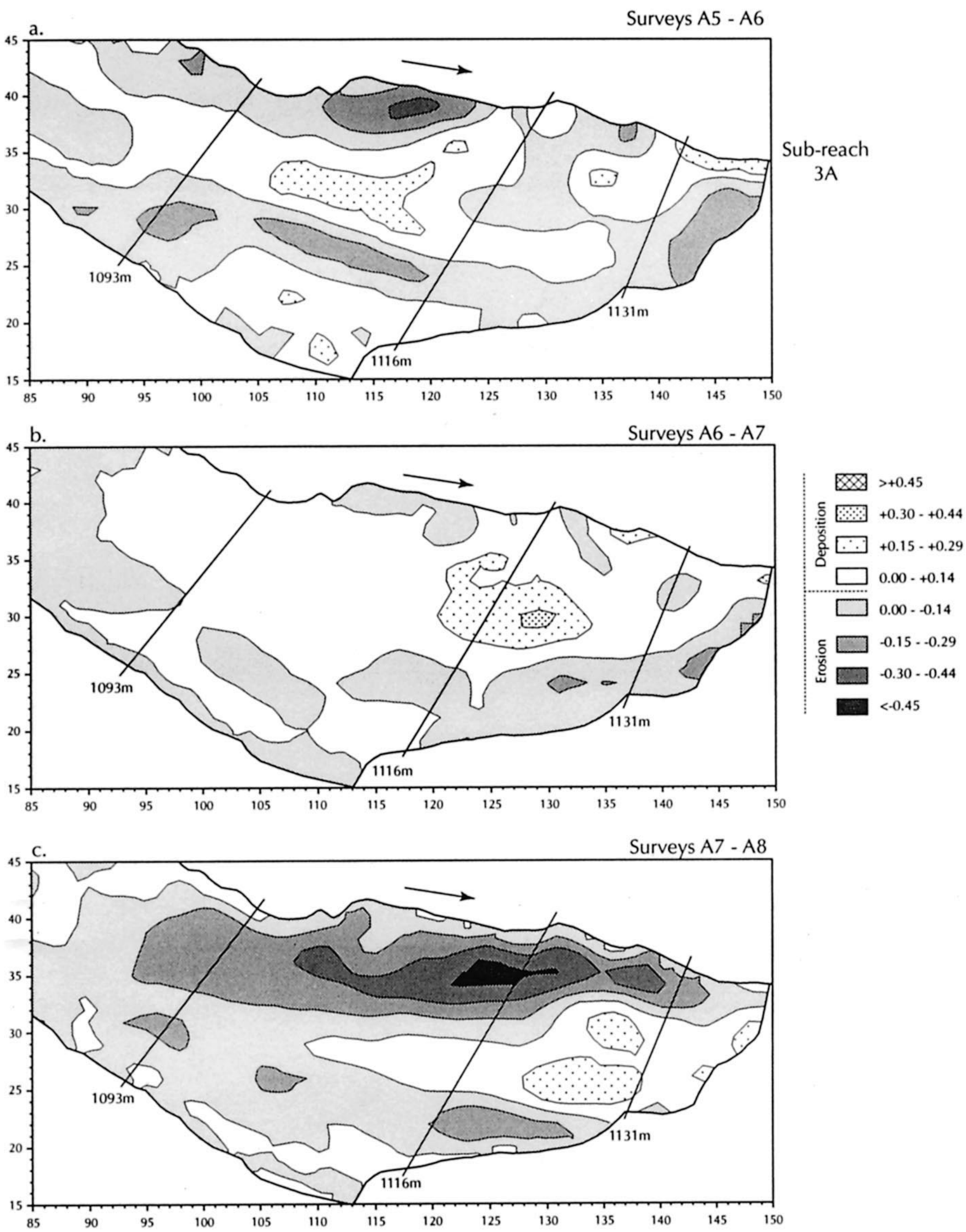

Figure 8. Spatial distribution of deposition and erosion in 3A between surveys A5 and A6 (a), A6 and A7 (b) and A7 and A8 (c). Note initial aggradation and further erosion of the medial bar (a) and substantial erosion (c) in response to previous aggradation as sediment was transferred into 3A from 1A and 2A (b). The sub-reach boundary excludes the abandoned channel to the right of the medial bar (see Figure 2) which was inactive during the study. Units are metres. Labelled cross-sections enable positions to be located with reference to Figures 1 and 2. Arrows show flow direction.

(e.g. Meade, 1985) on larger rivers. Passage of these megascale sediment waves can cause considerable aggradation (e.g. up to 5 m (Knighton, 1989; James, 1991); up to 9 m (Madej and Ozaki, 1996)) and planform change (Knighton (1989) reported a 300 per cent increase in channel width).

As the wave passed through Reach A, sediment was exchanged with all the sub-reaches. Sediment was deposited on forced bars in the stable upstream transfer sub-reaches (previous surveys indicate minimal changes here) causing local instability which returned some different sediment to the wave. This sediment was

supplied to the over-aggraded 3A which increased the wave magnitude as sediment exchange from the wave to the reach resulted in further instability causing a larger amount of different sediment to be transferred from the reach to the wave. The study duration and incomplete downstream dataset prevent quantification of the overall effect of Reach A upon wave attenuation. However, it is clear from these results that attenuation depends upon the interaction between the wave sediment and reach morphology. For example, if a storage reach is starved of sediment then attenuation of a passing wave would increase, whereas if the reach is 'over-aggraded' the wave may trigger instability and attentuation would be decreased (e.g. 3A), and may (as here) be replaced by amplification.

The wave sediment delivered from Reach A was not the same as the sediment delivered to the start of the reach, reflecting a complex chain of events within the reach. Deposition at some locations caused erosion at others with the resultant sediment being supplied to downstream reaches where the process was repeated. Sediment exchange between the reach and the wave controls the passage of this macroscale sediment wave with this interaction depending on the amount of incoming sediment, the recipient morphology and the channel response to local aggradation. The movement of magnetic tracer particles shows that coarser sediment entering with the wave remains in storage for longer than finer (*c.* D_{50}) material (Wathen, 1995). Total tracer numbers are inadequate for calculation of exchanges between the wave sediment and that in prior storage within Reach A. These controls differ from those which affect previously reported larger scale waves (e.g. Meade, 1985; Knighton, 1989) where there is little exchange between wave material and recipient morphology as the quantity and calibre of wave sediment overwhelms the natural transport processes within the recipient channel. Channel morphology is completely 'drowned out' by the incoming material and the channel attains a new state as hydraulics, sedimentology and planform all respond (e.g. Gilbert, 1917; Beschta, 1983; Pickup *et al.*, 1983; Meade, 1985; Knighton, 1989; Madej and Ozaki, 1996). This interaction is controlled by the magnitude, duration and calibre of sediment supply from the disturbance and the magnitude of recovery processes (e.g. Anderson and Calver, 1977; Wolman and Gerson, 1978).

Macroscale sediment waves are a significant control upon the behaviour of braided channels (Hoey, 1992; Goff and Ashmore, 1994) and reflect the interaction between discharge, morphology and sediment delivery (e.g. Lane *et al*., 1996). Local sediment supply from a wave can cause bar development and subsequent scour or avulsion may release a new sediment wave (e.g. Ashmore, 1991), possibly with increased magnitude (Hoey and Sutherland, 1991). The results from this study indicate that macroscale waves are also a significant control upon the morphology and behaviour of active single-thread channels. Whether a wave input will alter morphology and release new wave sediment depends upon sediment delivery and the stability of the recipient morphology; this is dependent upon previous sediment supply including the impact of previous waves. Macroscale sediment waves are controlled by the geomorphological regime of the channel (Hoey, 1992); however, they also partly control this regime. The permanence of the effect of a landform becomes a determining factor for future processes (Brunsden, 1990). If the impact of a wave persists until the next extreme event (3A may provide an example of this), then the possibility for renewed channel change in response to an incoming wave (and release of new wave sediment, e.g. 3A) is enhanced. If a wave does not enter, an extreme event may trigger channel change and release a new sediment wave. During normal flows, restorative processes act to return the channel towards equilibrium in these zones; however, if the timescale for recovery is shorter than the recurrence interval of extreme events (e.g. Brunsden and Thornes, 1979) then wave generation or an extreme response to an incoming wave (e.g. 3A) will occur. These unstable zones are fundamental controls upon wave development and channel behaviour and, in turn, reflect the impact of previous waves. The existence of these unstable areas and their behaviour in response to sediment inputs indicates that exogenous macroscale sediment waves may be inherent controls upon the morphology and stability of active single-thread channels, particularly where their influence is preserved until the next wave enters a reach. Long-term detailed monitoring of active single-thread channels is required to assess the validity of this.

Church and Jones (1982) suggested that a river will develop a stepped profile in response to the input of a slug of sediment. However, they ignored reach morphology and slope was seen as the major control. This study has demonstrated that a one-dimensional approach (e.g. Giffiths, 1979; Weir, 1983; Pickup *et al.*, 1983; Needham and Hey, 1992) is an imperfect approximation as downstream changes to a wave depend upon three-dimensional sediment exchange processes which are a function of morphology (e.g. Hoey, 1992). Nicholas *et*

al. (1995) have demonstrated the significance of valley width variations for modelled sediment storage, and have qualitatively confirmed field observations (Church, 1983; Macklin and Lewin, 1989; Hoey, 1994). Future modelling strategies must accommodate downstream and lateral dispersion of wave sediment (Nicholas *et al*., 1995) in response to recipient morphology and predict the resultant release of new wave sediment. Given the importance of exchange and morphology, it is suggested that modelling of sediment exchange between small storage units (e.g. Kelsey *et al*., 1987) as a wave passes through might yield insights into the exchange mechanisms, as well as quantitative predictions of storage volumes.

CONCLUSIONS

The behaviour of macroscale endogenous sediment waves is rarely documented and, consequently, the mechanisms of their downstream transfer are incompletely understood. Most studies detail the passage of megascale waves where the incoming sediment overwhelms the recipient channel and the morphology is drowned out (e.g. Meade, 1985; Knighton, 1989). Macroscale endogenous waves interact with the reach(es) through which they pass with morphology and reach stability controlling the magnitude of this interaction.

The sediment wave described in this study was generated by an avulsion in response to an extreme snowmelt flood. The resultant wave entered Reach A, 150 m downstream of the generation area, and the passage of this wave from initial input to amplified output may be explained by a series of processes.

(1) Local aggradation in stable locations: a large proportion of sediment was deposited on forced bars in the upstream stable transfer reaches (sub-reaches 1A and 2A).
(2) Downstream transfer of wave sediment, much of which did not enter the reach as part of the wave: bar aggradation and flow confinement resulted in erosion from adjacent and downstream channels and trimming of bar margins, which transferred different sediment (considered as part of the wave) downstream.
(3) Local aggradation in unstable locations: sediment was delivered to 3A, an unstable reach which had degraded for 16 months prior to wave entry, resulting in aggradation.
(4) Morphological change in response to 'over-aggradation': increased storage in 3A enhanced local instability causing erosion and an increase in wave magnitude as it passed through.

These results indicate that there is considerable interaction between a macroscale sediment wave and the reach through which it passes, and that the interaction is a function of sediment input, reach morpholgy and local stability in response to aggradation. The magnitude of the processes observed in this study suggests that macroscale sediment waves are important controls on channel stability and migration in active single-thread channels.

Wave sediment which enters a reach does not necessarily correspond with the sediment which exits the reach as considerable sediment exchange between the wave and the channel may alter the form and magnitude of the wave. The rate of exchange depends upon the morphological stability of the deposition location of the wave sediment. If sediment is deposited in stable locations then it is unlikely to be released, but this may trigger erosion in adjacent areas (e.g. 1A and 2A). If sediment is deposited in unstable locations such as 3A, then the input may trigger widespread instability and increase the amount of sediment transferred as the wave (depending upon the existing storage potential of the recipient reach). This three-dimensional interaction between a wave and reach is dependent upon recipient morphology and is not considered by one-dimensional models of wave movement.

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REFERENCES

Anderson, M, G. and Calver, A. 1977. 'On the persistence of landscape features formed by a large flood', *Transactions of the Institute of British Geographers,* **2**, 243–255.

Ashmore, P. E. 1991. 'Channel morphology and bedload pulses in braided gravel-bed streams', *Geografiska Annaler,* **73A**, 37–52.

Ashmore, P. and Church, M. in press. 'Sediment transport and river morphology: a paradigm for study', in Klingeman, P. *et al*. (Eds), *Gravel-bed Rivers in the Environment,* Proceedings of the 4th International Workshop on Gravel-bed Rivers, Water Resource Publications, Colorado.

Beschta, R. L. 1983. 'Channel changes following storm induced hillslope erosion in the Upper Kowai Basin, Torlesse Range, New Zealand', *Journal of Hydrology (New Zealand),* **22**, 91–111.

Brunsden, D. 1990. 'Tablets of stone: towards the Ten Commandments of geomorphology', *Zeitschrift für Geomorphologie Supplement,* **79**, 1–37.

Brunsden, D. and Thornes, J. B. 1979. 'Landscape sensitivity and change', *Transactions of the Institute of British Geographers,* **4**, 485–515.

Church, M. 1983. 'Patterns of instability in a wandering gravel-bed channel', in Collinson, J. D. and Lewin, J. (Eds), *Modern and Ancient Fluvial Systems,* International Association of Sedimentologists Special Publication, **6**, 169–180.

Church, M. and Jones, D. 1982. 'Channel bars in gravel bed rivers', in Hey, R. D., Bathurst, J. C. and Thorne, C. R. (Eds), *Gravel Bed Rivers: Fluvial Process, Engineering and Management,* Wiley, Chichester, 291–339.

Deitrich, W. E., Dunne, T., Humphrey, N. F. and Reid, L. M. 1982. 'Construction of sediment budgets for drainage basins', in Swanson, F. J., Janda, R. J., Dunne, T. and Swanston, D. N. (Eds), '*Sediment Budgets and Routing in Forested Drainage Basins*', USDA Forest Service, Pacific NW Forest and Range Experiment Station, General Technical Report **PNW-141**, 165 pp.

Ferguson, R. and Ashworth, P. 1991. 'Slope-induced changes in channel character along a gravel-bed stream: the Allt Dubhaig, Scotland', *Earth Surface Processes and Landforms,* **16**, 65–82.

Ferguson, R. I. and Ashworth, P. J. 1992. 'Spatial patterns of bedload transport and channel change in braided and near braided rivers', in Billi, P., Hey, R. D., Thorne, C. R. and Tacconi, P. (Eds), *Dynamics of Gravel Bed Rivers*, Wiley, Chichester, 477–496.

Ferguson, R. I. and Werritty, A. 1983. 'Bar development and channel changes in the gravely River Feshie, Scotland', in Collinson, J. D. and Lewin, J. (Eds), *Modern and Ancient Fluvial Systems*, International Association of Sedimentologists Special Publication, **6**, 181–193.

Ferguson, R. I., Hoey, T. B., Wathen, S. J. and Werritty, A. 1996. 'Field evidence for rapid downstream fining of river gravels through selective transport', *Geology,* **24**, 179–182.

Gilbert, G. K. 1917. *Hydraulic mining debris in the Sierra Nevada,* US Geological Survey Professional Paper, **105**, 154 pp.

Goff, J. R. and Ashmore, P. 1994. 'Gravel transport and morphological change in braided Sunwapta river, Alberta, Canada', *Earth Surface Processes and Landforms,* **19**, 195–213.

Griffiths, G. A. 1979. 'Recent sedimentation history of the Waimakariri River, New Zealand', *Journal of Hydrology (New Zealand),* **18**, 6–28.

Hassan, M. A., Church, M. and Schick, A. P. 1991. 'Distance of movement of coarse particles in gravel bed streams', *Water Resources Research,* **27**, 503–511.

Hoey, T. B. 1992. 'Temporal variations in bedload transport rates and sediment storage in gravel-bed rivers', *Progress in Physical Geography,* **16**, 319–338.

Hoey, T. B. 1994. 'Patterns of sediment storage in the Kowai River, Torlesse Range, New Zealand', *Journal of Hydrology (New Zealand),* **32**, 1–15.

Hoey, T. B. 1996. 'Sediment dispersion and duration of storage in a model braided river', *Journal of Hydrology (New Zealand),* **35**, 213–238.

Hoey, T. B. and Sutherland, A. J. 1991. 'Channel morphology and bedload pulses in braided rivers: a laboratory study', *Earth Surface Processes and Landforms,* **16**, 447–462.

James, L. A. 1989. 'Sustained storage and transport of hydraulic gold mining sediment in the Bear River, California', *Annals of the Association of American Geographers,* **79**, 570–592.

James, L. A. 1991. 'Incision and morphological evolution of an alluvial channel recovering from hydraulic mining sediment', *Geological Society of America Bulletin,* **103**, 723–736.

Kelsey, H. M., Lamberson, R. and Madej, M. A. 1987. 'Stochastic model for the long term transport of stored sediment in a river channel', *Water Resources Research,* **23**, 1738–1750.

Knighton, A. D. 1989. 'River adjustment to changes in sediment load: the effects of tin mining on the Ringarooma River, Tasmania, 1875–1984', *Earth Surface Processes and Landforms,* **14**, 333–369.

Kondolf, G. M. and Matthews, W. V. G. 1986. 'Transport of tracer gravels on a coastal California river', *Journal of Hydrology,* **85**, 265–280.

Kuhnle, R. A. and Southard, J. B. 1988. 'Bedload transport fluctuations in a gravel bed laboratory channel', *Water Resources Research,* **24**, 247–260.

Lane, S. N., Richards, K. S. and Chandler, J. H. 1994. 'Developments in monitoring and modelling small-scale river bed topography', *Earth Surface Processes and Landforms,* **19**, 349–369.

Lane, S. N., Richards, K. S. and Chandler, J. H. 1996. 'Discharge and sediment supply controls on erosion and deposition in a dynamic alluvial channel', *Geomorphology,* **15**, 1–15.

Macklin, M. G. and Lewin, J. 1989. 'Sediment transfer and transformation of an alluvial valley floor: the River South Tyne, Northumbria, U.K.', *Earth Surface Processes and Landforms,* **14**, 233–246.

Madej, M. A. and Ozaki, V. 1996. 'Channel response to sediment wave propagation and movement, Redwood Creek, California, USA', *Earth Surface Processes and Landforms,* **21**, 911–929.
McLean, D. G. 1990., *The relation between channel instability and sediment transport on Lower Fraser River*, unpublished PhD thesis, University of British Columbia.
Meade, R. H. 1985. 'Wavelike movement of bedload sediment, East Fork River, Wyoming', *Environmental Geology and Water Science,* **7**, 215–225.
Mosley, M. P. 1978. 'Bed material transport in the Tamaki River near Dannevirke, North Island, New Zealand', *New Zealand Journal of Science,* **21**, 165–184.
Needham, D. J. and Hey, R. D. 1992. 'Dynamic modelling of bed waves', in Billi, P., Hey, R. D., Thorne, C. R. and Tacconi, P. (Eds), *Dynamics of Gravel Bed Rivers,* Wiley, Chichester, 401–414.
Nicholas, A. P., Ashworth, P. J., Kirkby, M. J., Macklin, M. G. and Murray, T. 1995. 'Sediment slugs: large scale fluctuations in fluvial sediment transport rates and storage volumes', *Progress in Physical Geography,* **19**, 500–519.
Pickup, G., Higgins, R. J. and Grant, I. 1983. 'Modelling sediment transport as a moving wave – the transfer and deposition of mining waste', *Journal of Hydrology,* **60**, 281–301.
Seminara, G. and Tubino, M. 1989. 'Alternate bars and meandering', in Ikeda, S. and Parker, G. (Eds), *River Meandering*, Water Resource Monograph Series, **12**, AGU, Washington DC, 267–320.
Warburton, J. and Davies, T. 1994. 'Variability of bedload transport and channel morphology in a braided river hydraulic model', *Earth Surface Processes and Landforms,* **19**, 403–421.
Wathen, S. J. 1995. *The effect of storage upon sediment transfer processes in a small Scottish gravel-bed river*, unpublished PhD thesis, University of St Andrews (2 vols).
Wathen, S. J., Ferguson, R. I., Hoey, T. B. and Werritty, A. 1995. 'Unequal mobility of sand and gravel in weakly bimodal sediment', *Water Resources Research,* **31**, 2087–2096.
Weir, G. J. 1983. 'One dimensional bed wave movement in lowland rivers', *Water Resources Research,* **19**, 627–663.
Wolman, M. G. and Gerson, R. 1978. 'Relative scales of time and effectiveness in watershed geomorphology', *Earth Surface Processes,* **3**, 189–208.

